

# Climate and ozone response to increased stratospheric water vapor

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## Abstract

Stratospheric water vapor abundance affects ozone, surface climate, and stratospheric temperatures. From 30–50 km altitude, temperatures show global decreases of 3–6 K over recent decades. These may be a proxy for water vapor increases, as the GISS climate model reproduces these trends only when stratospheric water vapor is allowed to increase. Observations suggest that stratospheric water vapor is indeed increasing, though measurements are extremely limited in either spatial coverage or duration. Model results suggest that the observed changes may be part of a global, long-term trend.

Furthermore, the required water vapor change cannot be accounted for by increased stratospheric production, suggesting that climate change may be altering tropospheric input. The calculated water vapor increase contributes an additional  $\approx 24\%$  ( $\approx 0.2 \text{ W/m}^2$ ) to the global warming from well-mixed greenhouse gases over the past two decades. Observed ozone depletion is also better reproduced when destruction due to increased water vapor is included. If the trend continues, it could increase future global warming and impede stratospheric ozone recovery.

## Introduction

Observations at 30–50 km from satellites, rocketsondes and lidars show 1–2 K/decade cooling over the two to three decades of data [as in *World Meteorological Organization, 1999* (hereafter WMO99)]. Several factors have likely influenced these trends. Observations show decreases in upper stratospheric ozone. In percentage terms, these have been largest at 40–50 km altitude ( $\approx 2$ –1 mb), and at 15–20 km ( $\approx 150$ –70 mb), where they are 6–10%/decade [WMO99]. This depletion will have cooled these regions [Ramaswamy *et al.*, 1996]. Concurrently, the well-mixed greenhouse gases (ghgs), CO<sub>2</sub>, CH<sub>4</sub>, N<sub>2</sub>O, and the chlorofluorocarbons have increased, also cooling the stratosphere.

Stratospheric water vapor changes would have a similar effect [Rind and Lonergan, 1995; Forster and Shine, 1997, 1999; WMO99]. Its sources are *in situ* production from methane oxidation, and transport from the troposphere. The latter is governed largely by tropical tropopause temperatures [*e. g.* Mote *et al.*, 1996], and could be greatly altered by changes as small as a few tenths of a degree [Evans *et al.*, 1998]. Cross-tropopause transport can also be limited by water vapor availability. An increase in tropospheric water vapor, predicted by most climate models, could therefore lead to increased stratospheric water even without a warmer tropopause. Though the observational record is limited, seven years of satellite (Halogen Occultation Experiment - HALOE) and ground-based data for the middle and upper stratosphere [Evans *et al.*, 1998; Nedoluha *et al.*, 1998; Randel *et al.*, 1999] and 14 years of lower stratospheric measurements over Boulder, CO [Oltmans and Hofmann, 1995] show large increases in stratospheric water vapor. Aircraft observations, however, give conflicting indications

of trends in the lower stratospheric [Peter, 1998; Hurst *et al.*, 1999]. Given these uncertainties, the question remains open as to current and future trends in stratospheric water vapor.

## Model Description

We have used the Goddard Institute for Space Studies (GISS) general circulation model (GCM) [*e.g.* Rind *et al.*, 1998] to investigate long-term upper stratospheric cooling and ozone depletion. The version used has 8° x 10° resolution, and extends from the surface to 85 km. At this high altitude allows for a good simulation of middle and upper stratospheric temperatures and circulation. The model was run with increasing chlorine and GHGs based on observations through the 1980s, thereafter using projected trends [Shindell *et al.*, 1998a]. To calculate ozone amounts, parameterized chemistry derived previously from our 2D photochemical model [Shindell *et al.*, 1998b] was installed in the GCM. Ozone responds to constituent anomalies relative to 1979, along with temperature changes and changes in local radiative flux due to differences in overlying ozone. Polar stratospheric clouds (PSCs) do not respond to water vapor changes, however, and ozone transport does not interact with the radiation, though these are important primarily in the lower stratosphere. While the chemistry is simplified, use of a full GCM allows us to simulate changes in tropical tropopause temperatures and in circulation, and to assess the climate response.

Four experimental runs were performed: GHGs, with changes in the well-mixed ghgs, but fixed water and ozone (G); Ozone, with ghg and chlorine changes, calculated ozone, and fixed water vapor (G + O); MethOx, with ghg and chlorine changes, calculated ozone,

and water vapor increases due to methane oxidation, as derived from the 2D model (G + O + M); and Water, with ghg and chlorine changes, calculated ozone, and the GCM's internally calculated water vapor plus increases from methane oxidation (G + O + M + W). In the MethOx and Water runs, water is allowed to change throughout the stratosphere and ozone is allowed to respond. Model increases in the lower stratosphere lie between balloon and satellite observations (Table 1). The GCM values generally follow the vertical shape of the observed trends, but have smaller magnitudes than the HALOE data.

## Simulated Trends

The model reproduces the overall spatial distribution and magnitude of the observed ozone trends (Figure 1). The upper stratosphere is of greatest interest here, and where the observations are statistically significant. Agreement there is best when water vapor increases are included. Ozone trends are less certain in the lowermost stratosphere [WMO99]. Simulated decreases there are near the low end of trend estimates, as for most photochemical models [Dvortsov *et al.*, 1999], but inclusion of water induced destruction again improves the response. The net ozone change results from several mechanisms (Figure 2). Increased chlorine dominates upper stratospheric ozone depletion, with an additional contribution from water vapor. These are partially offset by ghg induced cooling, which slows ozone destroying chemical reactions, and by the direct chemical effect of increasing methane (not shown), which removes chlorine from forms that destroy ozone [*e. g.* Siskind *et al.*, 1998].

We compare modeled and observed temperature trends at 43 N, where measurements

are dense (Table 2). The model's upper stratosphere cools in response to increasing ghgs and ozone depletion, but much less than the observed trends at 40 km and above. Including a water vapor increase due to methane oxidation and the ozone response improves the result, but only with the inclusion of additional water vapor can the model reproduce the observed trends within their  $1\sigma$  uncertainty. Similar results were found at other latitudes where observations were available (see also Table 2 caption).

While the temperature response to ghgs in the GISS model agrees well with other GCMs (e.g. WMO99, p. 12.30, showing the responses to increasing ghgs and halogens, similar to the Ozone run), it differs from fixed dynamical heating (FDH) models [Forster and Shine, 1999; Ramaswamy *et al.*, 1992]. As shown in Figure 2 of [Rind *et al.*, 1998], the change in temperature by dynamics becomes important at and above about 50 km. The heating change by dynamics in the present experiments was about 0.2 K at 50 km. The maximum cooling from ghg increases thus occurs at lower altitudes in the GCM than in the simpler FDH models.

Temperature trends in the Ozone, MethOx and Water runs also depend upon the water vapor and ozone changes. Our values maximize in the upper stratosphere, where the water vapor increases fastest (see Tables). In contrast, Forster and Shine's work included a uniform water increase throughout the stratosphere, at values appropriate for the lower stratosphere, their primary interest. Our temperature response in the upper stratosphere is therefore larger than theirs. Additionally, details of the radiative transfer calculation are important, with codes using fine resolution showing cooling from water more evenly distributed with altitudes than

broadband codes.

Caution must be exercised in comparing these simulations with those of models using observed ozone changes. The latter cannot separate out the total impact of water on temperature trends, as the ozone changes include chemical effects of water on ozone. This is especially important in the lower stratosphere and above about 5 mb (Fig. 2).

Long-term monitoring of stratospheric water vapor will be necessary to definitively establish trends. However, these results suggest that stratospheric water must have increased over the past several decades. Furthermore, the increase was larger than that due to methane oxidation alone, as in recent measurements [Evans *et al.*, 1998; Nedoluha *et al.*, 1998; Oltmans and Hofmann, 1995]. Thus a portion of the increase resulted from enhanced transport from the troposphere. This occurs in the GCM due to warming of the tropical tropopause by increasing ghgs. Observations do not indicate an overall warming trend at the tropical tropopause [Simmons *et al.*, 1999]. Given the complexity of cross-tropopause transport of water vapor, however, this does not rule out an increased flux of water vapor into the stratosphere due to climate change.

Temperature trends vary markedly between simulations, especially in the future. The GHG run shows a steady cooling at 0.7 mb ( $\approx 50$  km) (Figure 3). Including ozone depletion, the trend accelerates from 1959 to about 2000, then levels off as ghg induced cooling is offset by warming from ozone recovery. Thus without increases in water, the rapid upper stratospheric cooling should soon level out. However, with increases, there should be continuation of this cooling. Behavior at other altitudes is similar, with larger effects with increasing altitude.

Future ozone trends also show large differences. At high altitudes, increased water decreases ozone (Fig. 2). In the middle stratosphere, ozone is less sensitive to hydrogen oxides, so there is a net increase from water-induced cooling and increased radiation due to less overhead ozone, as in Evans *et al* [1998]. In the lower stratosphere, which dominates the total column, ozone is again quite sensitive to water. Nitrogen removal through heterogeneous reactions allows more ozone destruction by chlorine, so that currently, with abundant chlorine, water increases lead to a net chemical ozone loss. If future chlorine is greatly reduced, nitrogen removal should increase ozone by reducing ozone depleting reactions with nitrogen oxides. Increasing water vapor delays recovery of the ozone column past the peak chlorine loading of 2000–2005 (Figure 4). While increasing ghgs in the ozone run speed up ozone recovery, when water vapor increases are also included, recovery to 1979 levels at mid-latitudes is extended past the end of the simulations.

Given the simplified ozone anomaly model used here, along with difficulties in both defining and reproducing mid-latitude ozone trends near the tropopause, these future ozone column amounts should be treated as qualitative predictions of the influence of water vapor.

In the polar regions, “ozone holes” are created by massive ozone destruction resulting from chemical processing on the surfaces of PSCs. PSC formation could be enhanced by increased stratospheric water vapor [WMO99, Kirk-Davidoff *et al.*, 1999], leading to greater ozone losses and a further delay relative to that predicted due to ghg-induced cooling alone [Rind and Lonergan, 1995; Forster and Shine, 1997, 1999; Shindell *et al.*, 1988a; Dameris *et al.*, 1998].

## Climate Response and Conclusions

Changes in both stratospheric water vapor and ozone affect surface climate. From 1959 to 1999, simulated stratospheric ozone depletion caused a global-mean annual average surface air temperature change of  $-0.09$  K ( $-0.07$  K for 1979–1999). This is less than the  $-0.12$  to  $-0.20$  K calculated with the standard (limited stratosphere) GISS climate model using observed 1979–1994 ozone changes [Hansen *et al.*, 1997], due to the photochemical model’s underestimation of ozone losses in the lowermost stratosphere, where climate sensitivity to ozone change is largest. During 1959–1999, the water vapor increase in the Water run led to warming of  $+0.26$  K ( $+0.07$  K/decade). Forster and Shine [1999], who used the observed lower stratospheric water vapor trends, found a similar increase in surface air temperature:  $0.11$  K for the eighteen year period 1979–1997 ( $+0.06$  K/decade). The net change from stratospheric water and ozone was thus  $+0.17$  K, roughly 33% of the observed increase. Given large uncertainties in climate forcings from clouds and aerosols [Hansen *et al.*, 1998], such a large value is not inconsistent with current understanding.

These results suggests that impacts of stratospheric water vapor must be accounted for in climate and ozone studies. Changes in water and ozone are partially “forcings”, resulting directly from anthropogenic emissions, and partially “feedbacks”, resulting from the meteorological response to emission changes. A portion of water vapor trends is attributable directly to methane, while the remainder must be attributed to all the ghgs (including ozone) that affect circulation and tropopause temperature. For ozone changes, after another few decades, the influence of ghgs and water

vapor on ozone becomes larger than that of halogens, so that attribution should progressively switch from halogens to climate.

Future ozone recovery will affect its climate forcing. Ghg induced cooling maximizes in the upper stratosphere, leading to a relatively rapid recovery there. Furthermore, additional upper stratospheric ozone absorbs incoming radiation, reducing ozone production lower down. Since upper stratospheric ozone increases lead to surface cooling [Rind and Lacis, 1993], they largely offset the radiative impact of the slowly increasing abundances of lower stratospheric ozone. By 2050, the surface cooling due to ozone is reduced to  $-0.03$  K, while the change due to water vapor increases to  $+0.43$  K. The net stratospheric water-plus-ozone forcing may thus more than double over the coming half-century, contributing an additional 10–15% to the global warming due to well-mixed greenhouse gases.

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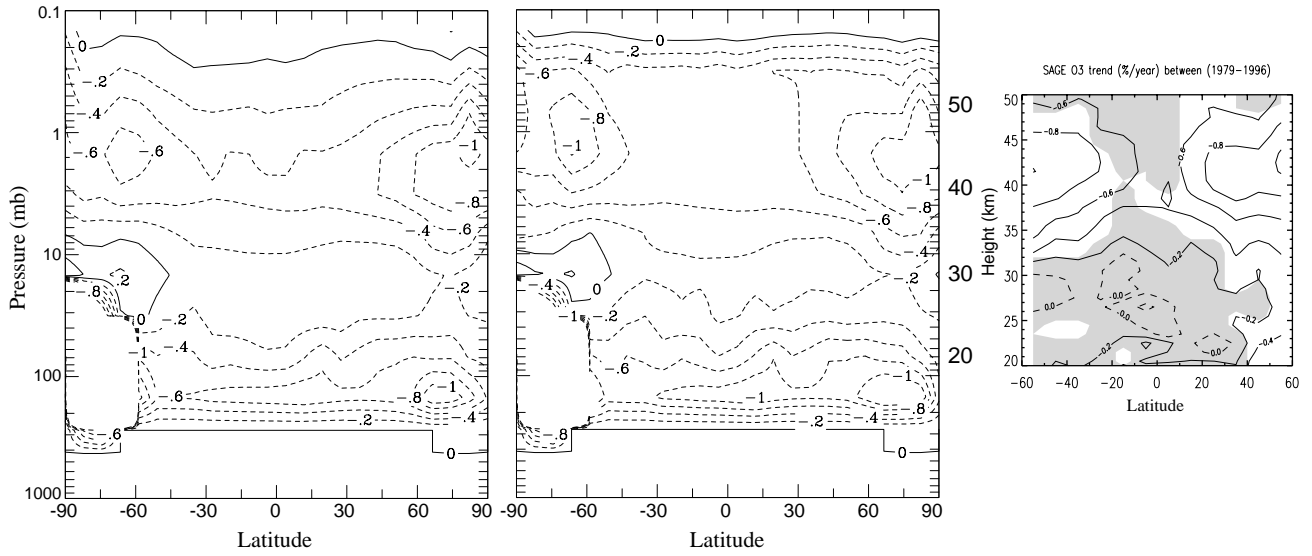
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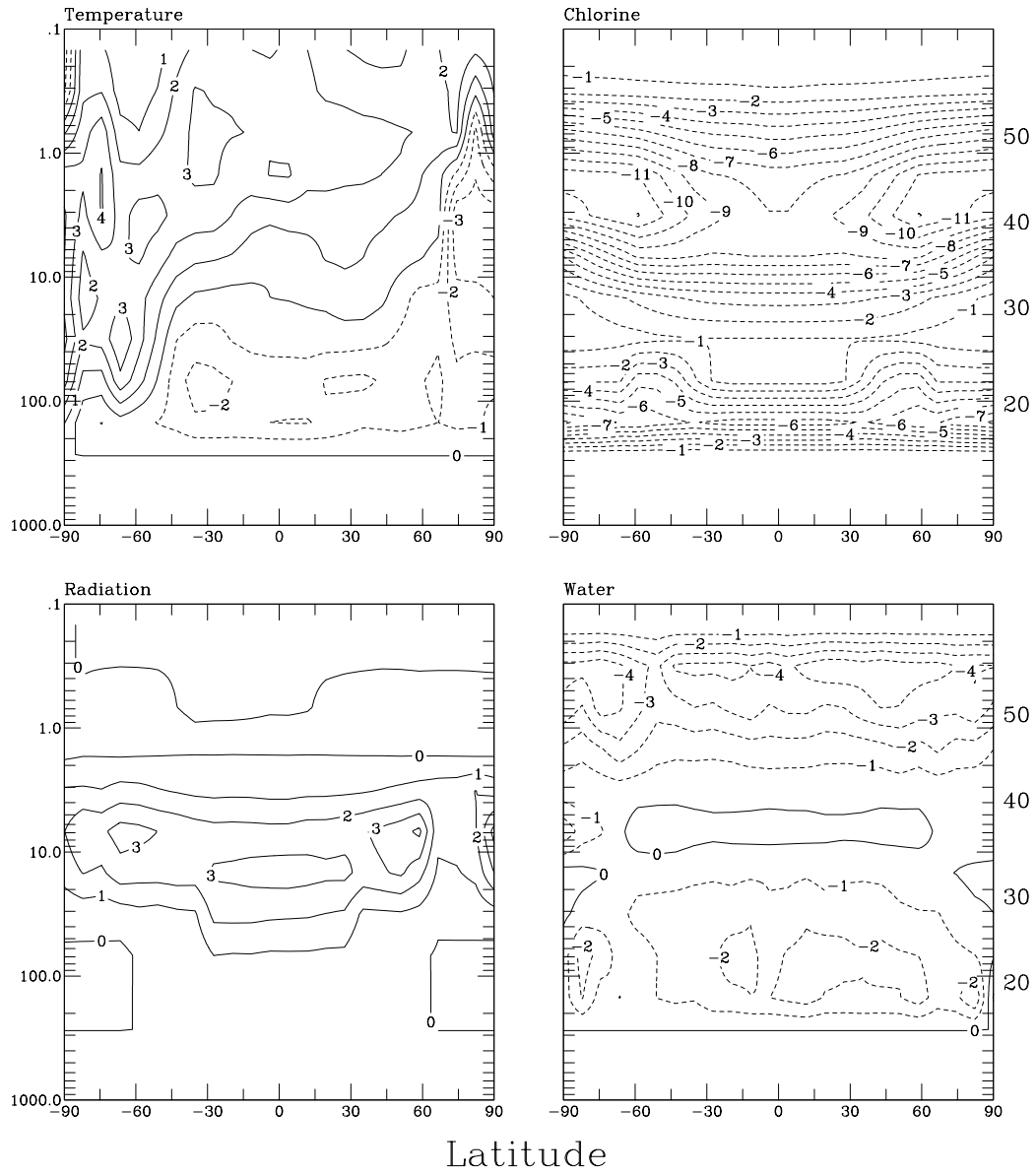
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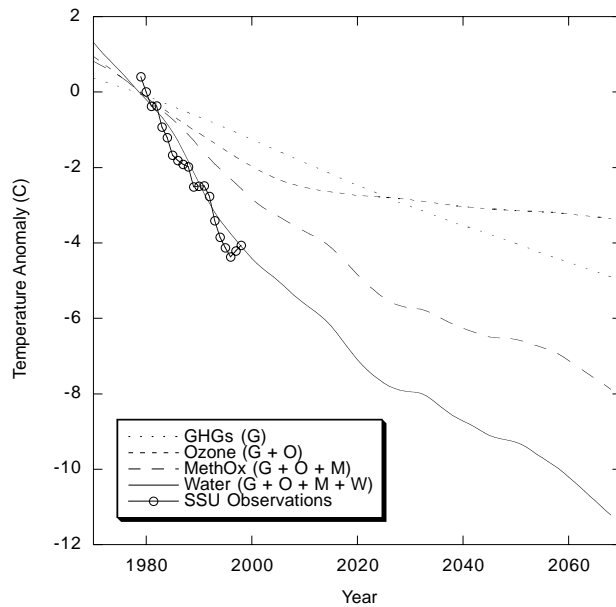


**Figure 1.** Modeled and observed annual average ozone trends from 1979 to 1996 in percent per year. Model results are shown for the Ozone (G + O) and Water (G + O + M + W) runs, in the left and center panels, respectively. Shading in the data from the Stratospheric Aerosol and Gas Experiment satellite (right) indicates no statistical significance. Note that those data only extend to about 60 degrees.

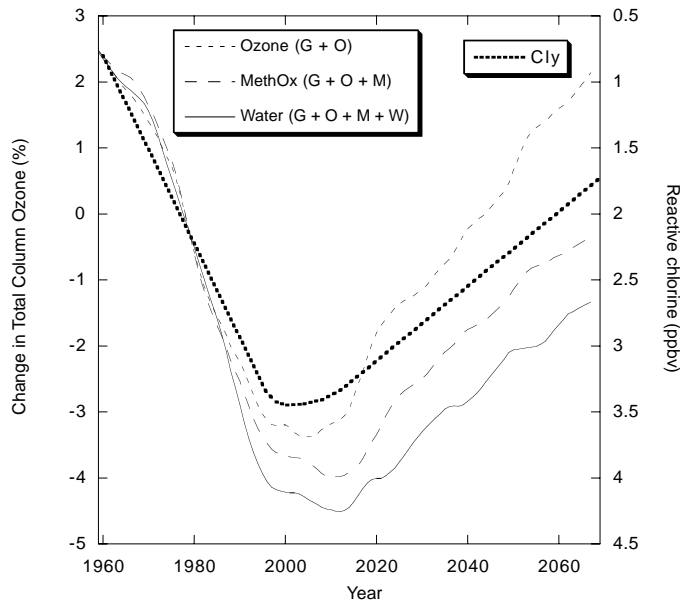




**Figure 2.** Annual average 1979 to 1996 trends in percent per decade from the MethOx run due to the indicated changes, based on emissions given in the text. Methane and nitrous oxide effects were smaller than those shown here. The chemical effect of water vapor is shown, which is partially canceled by its contribution to greenhouse cooling included under temperature.



**Figure 3.** Annually averaged temperature trends relative to 1980 over 60N–60S, at 0.7 mb ( $\approx 50$  km altitude). Model trends were smoothed with a ten-year weighting. Observations were taken by the Stratospheric Sounding Unit (SSU) over roughly 44–56 km altitude. There is little latitudinal structure to the modeled temperature changes, as in observations.



**Figure 4.** Column ozone trends at 43 N (left axis) and prescribed atmospheric chlorine loading (right axis). Ozone depletion is smoothed with a ten-year weighting. Results from the Ozone run are similar to predictions at northern mid-latitudes from 2D models. Observations from 1979 through 1997 show a decrease of about  $3.3 \pm 1.5$  percent per decade at this latitude, with indications of a slower loss rate since 1997 [WMO99]

**Table 1.** Water vapor trends (ppbv/yr)

Pressure (mb)	Altitude (km)	MethOx 43N	Water 43N	MethOx 60S-60N	Water 60S-60N	Balloons 40N	HALOE 60S-60N
15	29	19	40	20	51	15	60 - 65
6.8	34	20	44	21	57	na	85 - 90
3.2	39	22	61	21	66	na	90 - 95
1.5	44	24	75	23	78	na	110 - 115
0.7	50	27	67	26	69	na	100 - 105

Model altitudes are approximate values for the GCM layers centered at the given pressure levels. Balloon data are for 24–26 km altitude, the highest level reported [Oltmans and Hofmann, 1995], with uncertainty of  $\pm 15$  ppbv/yr at the 95 percent confidence level. HALOE data are fits to 1993–1997 measurements [Randel *et al.*, 1999], with uncertainties of approximately  $\pm 15$  ppbv/yr at 32 mb, and  $\pm 20$  ppbv/yr at 1 mb.

**Table 2.** Temperature change at 43 N, K/decade, 1979–1994

Pressure (mb)	Altitude (km)	Observations $\pm 1\sigma$	GHG run (G)	Ozone run (G+O)	MethOx run (G+O+M)	Water run (G+O+M+W)
15	29	$-0.80 \pm 0.29$	-0.24	-0.29	-0.40	<b>-0.53</b>
6.8	34	$-0.88 \pm 0.30$	-0.47	<b>-0.63</b>	<b>-0.73</b>	<b>-0.84</b>
3.2	39	$-1.23 \pm 0.33$	-0.46	-0.70	-0.84	<b>-1.43</b>
1.5	44	$-1.81 \pm 0.37$	-0.45	-0.74	-0.94	<b>-1.68</b>
0.7	50	$-2.55 \pm 0.40$	-0.33	-0.63	-1.37	<b>-2.32</b>

Altitudes for the GCM runs are approximate, as in Table 1. Observations used several techniques, with an average latitude of 45 N, and altitudes at 5 km intervals beginning at 30 km [WMO99]. Modeled values lying within  $1\sigma$  of the observations are shown in bold type. Latitudinal structure in the GCM does not alter the conclusion that water vapor increases are required to reproduce the observed trends. For comparison, the 60S-60N temperature trend averages for the GHG run are -0.33, -0.46, -0.55, -0.60, and -0.64 with increasing altitude, while for the Ozone run, they are -0.39, -0.60, -0.75, -0.85 and -0.92. See also Figure 3 for 60S-60N averages.